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Larry Benson
University of Colorado at Boulder, great.basin666@gmail.com

Richard Madole
US Geological Survey

William Phillips
University of Edinburgh

Gary Landis
US Geological Survey

Terry Thomas
New Mexico Bureau of Geology

See next page for additional authors

Authors
Larry Benson, Richard Madole, William Phillips, Gary Landis, Terry Thomas, and Peter Kubik

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The probable importance of snow and sediment shielding on cosmogenic ages of north-central Colorado Pinedale and pre-Pinedale moraines

Larry Benson[a,*], Richard Madole[b], William Phillips[c], Gary Landis[d], Terry Thomas[e], Peter Kubik[f]

[a] US Geological Survey, 3215 Marine Street, Boulder, CO 80303-1066, USA
[b] US Geological Survey, MS 980, Denver Federal Center, Lakewood, CO 80225, USA
[c] School of Earth, Environmental and Geographical Sciences, Drummond Street, University of Edinburgh, Scotland EH8 9XP, UK
[d] US Geological Survey, MS 963, Denver Federal Center, Lakewood, CO 80225, USA
[e] New Mexico Bureau of Geology, 801 Leroy Place, Socorro, NM 87801, USA
[f] Paul Scherrer HPK H30, ETH Institut of Particle Physics, Hönggerberg, Zurich CH-8093, Switzerland

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Abstract

Eight uncorrected ^{36}Cl ages for Pinedale boulders in north-central Colorado fall in the range 16.5 to 20.9 kyr. ^{10}Be age determinations on four of five boulders are in close agreement ($\leqslant$6% difference) with ^{36}Cl determinations. Hypothetical corrections for snow shielding increased the ^{36}Cl ages of Pinedale boulder surfaces by an average of ~12%. Most ages for pre-Pinedale (Bull Lake) boulders fall within marine-isotope stage (MIS) 5, a time when continental and Sierran ice accumulations were small or nonexistent. Under the assumption that these boulders were deposited on moraines that formed before the end of MIS 6 (~140 kyr BP), calculations indicated that rock-surface erosion rates would have had to range from 5.9 to 10.7 mm kyr^{-1} to produce the observed ^{36}Cl values. When compared to rates that have been documented for the past 20 kyr, these erosion rates are extremely high. Snow shielding accounts for 0–48% of the additional years needed to shift pre-Pinedale dates to MIS 6. This suggests that some combination of snow shielding, sediment shielding, or ^{36}Cl leakage has greatly decreased the apparent ages of most pre-Pinedale boulders. Inability to account for the effects of these processes seriously hinders the use of cosmogenic ages of pre-Pinedale boulders as estimators of the timing of alpine glaciation.
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1. Introduction

Cosmogenic dating has the potential for filling large gaps in our understanding of the late Quaternary glacial chronology of the Colorado Rockies. Prior to the application of cosmogenic dating, the timing of Pinedale glacial events was estimated from a few ^{14}C dates determined on the organic component of lake and fen sediments that accumulated in depressions on or adjacent to glacial deposits (Table 1). Most of these ^{14}C dates provide only minimum age estimates of glacial events. In addition, ^{14}C ages of organic material in till have not been reported for any locality in the north-central Rocky Mountains.

Radiocarbon ages of basal sediment in fens and ponds located upvalley from terminal moraines have not been extremely useful in reconstructing glacial histories. Most ^{14}C ages from these settings do not date glacial events closely, and estimating the amount of time elapsed between deposition of glacial sediment and deposition of overlying lake or fen sediment is invariably problematic. Therefore, cosmogenic dates may better constrain the glacial histories of the valleys.

The purpose of this paper is to present the results of ^{36}Cl and ^{10}Be dating of boulders on Pinedale and pre-Pinedale moraines in the Middle Boulder Creek, North St. Vrain, and Roaring Fork drainages (Figs. 1 and 2) of north-central Colorado, and to demonstrate the importance of snow and sediment shielding on the results.

*Corresponding author. Tel.: +1-303-541-3005; fax: +1-303-447-2505.
E-mail address: lbenson@usgs.gov (L. Benson).

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Table 1
Elevations and ^{14}C ages of samples used to approximate the advance and retreat of Pinedale glaciers in north-central Colorado

Site	Locality	Ele.(m)	Sample material	^{14}C age (kyr BP)	Lab number	Reference
Basal sediment from ice-marginal lakes located near Pinedale terminal moraine						
1	Devlin Park	2956	LOC	22.40 ± 1.23	DIC-870	Madole (1986)
2	Triangle Park	2990	LOC	23.17 ± 1.05	DIC-1170	Rosenbaum and Larson (1983)
Kettle pond (3) *and fen* (4) *located on Pinedale terminal moraine*						
3	Winding River	2640	LOC	13.82 ± 0.81	GAK-4537	Madole (1976b)
4	Mary Jane	2882	Peat	13.74 ± 0.16	DIC-671	Nelson et al. (1979)
Fen on Continental Divide site of major transect glacier						
5	La poudre Pass	3092	Peat	9.85 ± 0.30	W-4083	Madole (1980)
Ponds and fens located near center of Pinedale icecap						
6	Buffalo Pass 1	3156	LOC	13.68 ± 0.11	W-5028	Madole (1986)
7	Buffalo Pass 2	3168	LOC	11.76 ± 0.10	W-5034	Madole (unpublished)
Lakes near cirques						
8	Lake Dorothy	3677	HOC	10.91 ± 0.32	GX-17863	Davis et al. (1992)
9	Butterfly Lake	3460	LOC	10.41 ± 0.52	GX-11774	Davis (1987)
10	Blue Lake	3442	LOC	12.28 ± 0.34	GX-17878	Davis et al. (1992)
11	Sky Pond	3323	Beetle	12.04 ± 0.06	CAMS-22027	Menounos and Reasoner (1997)
12	Mitchell Lake	3262	LOC	10.92 ± 0.36	SI-5189	Nichols et al. (1984)
13	Long Lake	3208	LOC	11.80 ± 0.45	GX-7723	Nichols et al. (1984)

LOC and HOC refer to sediments containing low and relatively high amounts of organic carbon. See Fig. 1 for site locations.

2. Study area

Samples for cosmogenic dating were collected on the crests of end moraines near the downvalley limit of glaciation in three valleys in north-central Colorado. Two of the valleys are located on the east slope of the Front Range and one valley is located on the east slope of the Park Range (Fig. 1). Front Range samples are from the North St. Vrain Creek and Middle Boulder Creek (Figs. 2A and 2B) drainages, and Park Range samples are from the Roaring Fork valley (Fig. 2C). Proterozoic granite and quartz monzonite are the dominant bedrock types in the glacier accumulation area in all three valleys (Gable, 1969; Snyder, 1980; Braddock and Cole, 1990).

Most Pleistocene glaciers in north-central Colorado originated in cirques and flowed down deep valleys. In a few places, valley glaciers drained small icecaps, the largest of which mantled the summit of the Park Range east of Steamboat Springs (Madole, 1976a, 1980, 1991a, b). The glacier in the Cache la Poudre River valley (Fig. 1) was 42 km long and as much as 600 m thick, but most valley glaciers in the region were less than half as large (Madole et al., 1998).

The typical Front Range glacier attained lengths of 10–20 km and thicknesses of 180–350 m, and most terminated at elevations between 2500 and 2700 m. Valley glaciers in the Park Range were similar in length and thickness to those in the Front Range. However, glaciers on the east slope of the Park Range generally were only 10–13 km long because the relatively flat floor of a high intermontane basin (North Park) limited their eastward descent. Most glaciers on the west slope of the Park Range terminated at elevations between 2130 and 2200 m, which is much lower than their Front Range counterparts. Glaciers on the east slope of the Park Range terminated at elevations between 2530 and 2680 m (Madole 1991a, b). These glaciers tended to spread broadly and form sets of massive end moraines.

Many Front Range valleys are so deep and narrow that glaciers were unable to overtop valley walls and, consequently, lateral moraines are not well preserved. Most pre-Pinedale till in these valleys was overridden by Pinedale glaciers or was buried by valley-side colluvium. However, in those valleys where glacier termini were able to spread laterally, deposits of two, and in some places possibly three, different glaciations are preserved. In order of increasing age, the two widely recognized deposits are referred to as Pinedale and Bull Lake (Madole, 1976a; Madole et al., 1998). The relative ages of the deposits are based on differences in areal extent, stratigraphic position, topographic expression, and degree of weathering (Madole et al., 1998).

Pinedale deposits mantle 75–90% of the glaciated area. These deposits have a distinctive morphology that includes high (20–30 m), sharp-crested moraines, knob-and-kettle topography, and numerous fens and kettle ponds. The combined thickness of A and B horizons of soils in till of Pinedale age commonly ranges from as little as 50 cm near the downvalley limit of glaciation to

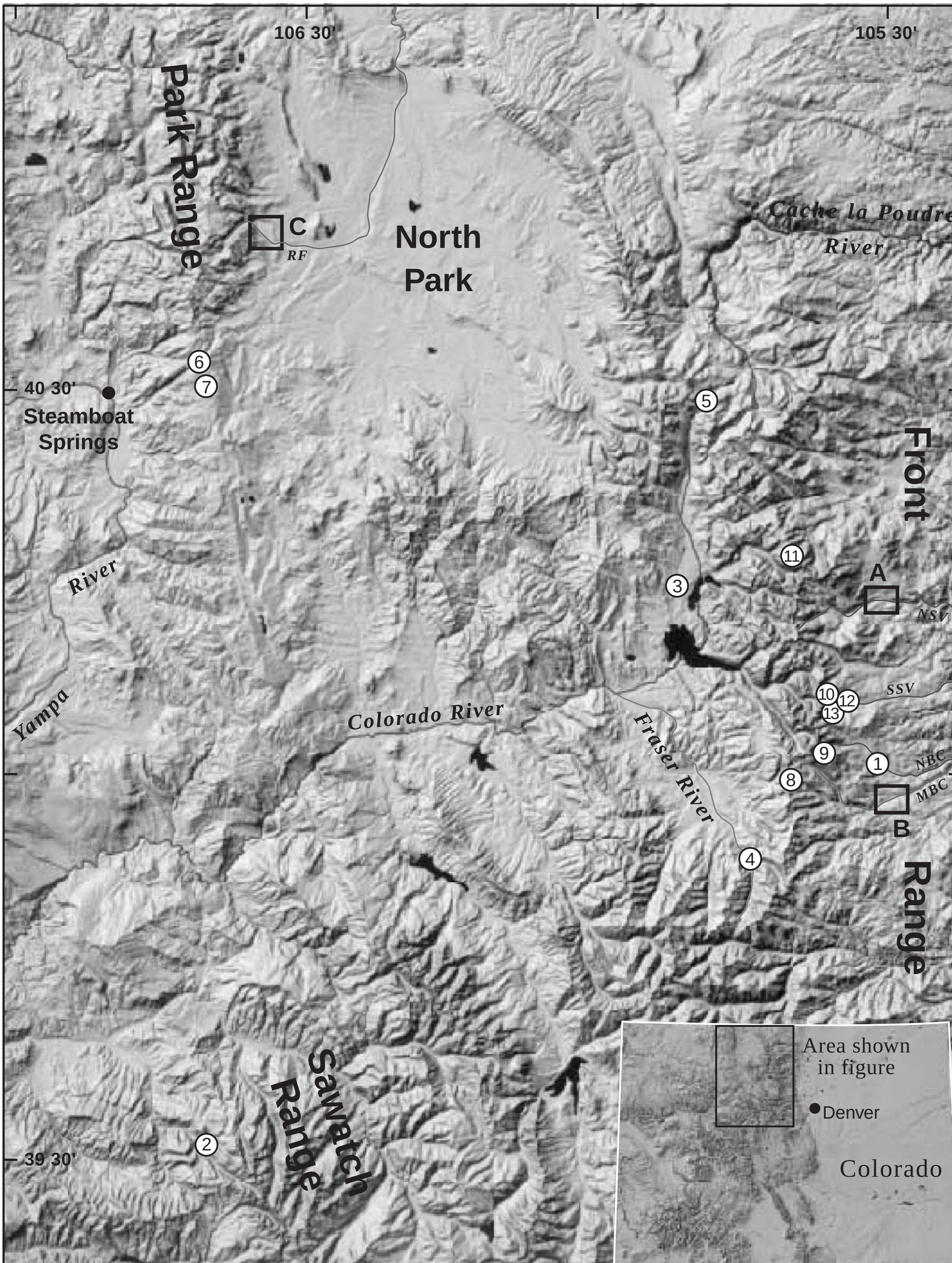

Fig. 1. Map showing locations of sample sites discussed in this paper. Circled numbers refer to sites listed in Table 1. Letters with boxes refer to locations of boulders sampled for cosmogenic dating in the North St. Vrain (NSV) drainage (A), Middle Boulder Creek (MBC) drainage (B), and Roaring Fork (RF) drainage (C). South St. Vrain (SSV) and North Boulder (NBC) creeks also shown.

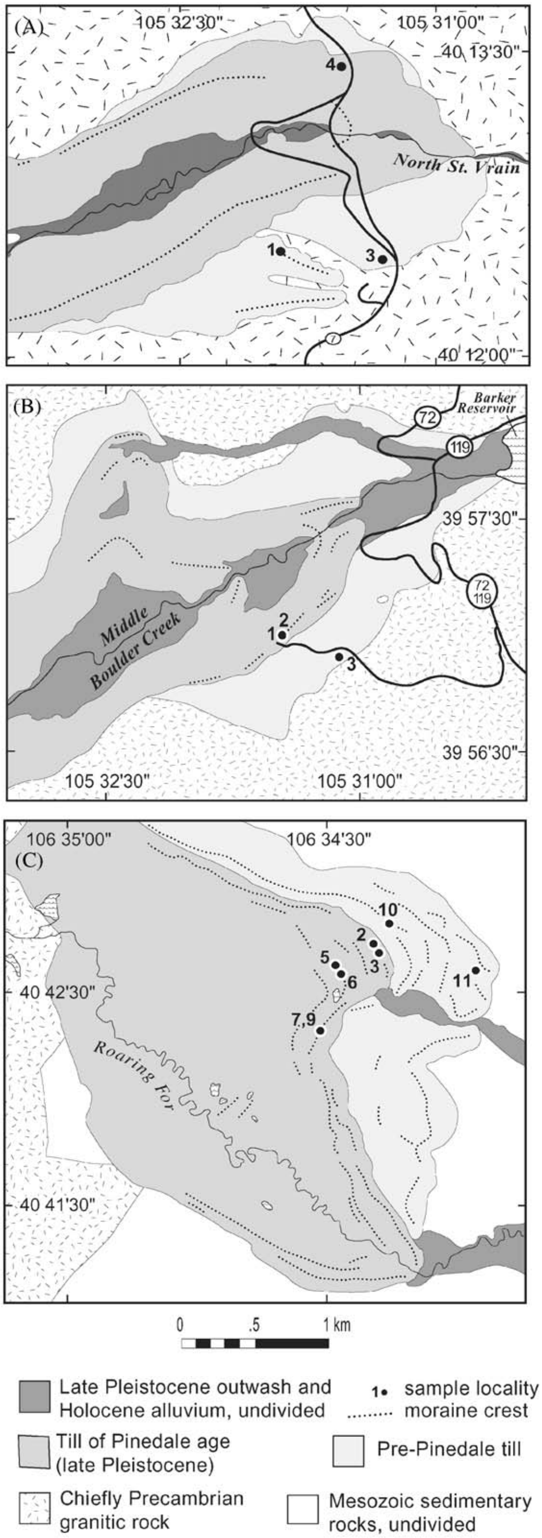

Fig. 2. Maps showing locations of boulders sampled for cosmogenic dating in the (A). North St. Vrain drainage, (B). Middle Boulder Creek drainage, and (C). Roaring Fork drainage.

as much as 100 cm at higher (>2900 m) elevations (Madole 1969; Madole and Shroba, 1979). Weathering rinds in most clasts are <3 mm thick and decomposed clasts are rare (Madole, 1969, 1976a). In contrast, end moraines of Bull Lake age are not as high or sharp-crested as are those of Pinedale age and, where preserved, knob-and-kettle topography is subdued and kettles rarely contain ponds. In addition, Bull Lake soils are much thicker (1.5 m) and more strongly developed than Pinedale soils, and 20–30% of the clasts in them are partly to completely decomposed (Madole, 1969, 1976a).

Although differences between deposits of Pinedale and Bull Lake age are pronounced, the same is not true for deposits older than Pinedale. Therefore, for the purposes of this paper, we will refer to Colorado glacial deposits that are older than Pinedale simply as pre-Pinedale deposits.

3. Radiocarbon-based estimates of the timing of Pinedale glaciation in north-central Colorado

In this section we discuss and summarize ^{14}C-based studies of Pinedale glaciation in north-central Colorado. The best estimate for the initiation of Pinedale glaciation comes from the Mary Jane site located in the Fraser River valley (Fig. 1, Table 1) (Nelson et al., 1979). A sequence of interbedded diamicton, peat, and lacustrine sediment was exposed in a pit ($15\times18\times12\,m^3$) excavated for the foundation of a ski-lift tower. A thin, discontinuous bed of lake sediment with an age of 30.1 ± 1.2 ^{14}C (35.1 cal)[1] kyr BP was found sandwiched between diamicton beds that may correspond to the initial Pinedale advance. However, given the restricted area of viewable outcrop, it is not possible to completely rule out a landslide origin for the basal diamicton.

Because the inception of ice-marginal lakes is directly linked to glacier movement, the ^{14}C ages of organic matter from basal sediments of such lakes tend to date glacial events more closely than the ^{14}C ages of organic matter from lake and fen sediments located in depressions on glacial deposits. Thus, ^{14}C dates from two ice-marginal paleolakes allow us to approximate the time of greatest expansion of Pinedale ice. Glacial Lake Devlin formed when the North Boulder Creek valley glacier blocked the mouth of a tributary located 2.3 km upvalley from the Pinedale terminal moraine. The inception of Lake Devlin, which occurred at or before 22.4 ± 1.1 ^{14}C (26.4 cal) kyr BP (Fig. 1, Table 1), provides a limiting date for the maximum expansion of Pinedale ice (Legg and Baker, 1980; Madole, 1986). Triangle

[1] ^{14}C ages between 9.0 and 12.0 kyr were converted to calendar (cal) yr using a second degree polynomial fit to the Cariaco Basin data of Hughen et al. (1998), and ^{14}C ages between 12.0 and 30.1 kyr were converted to calendar yr using a second degree polynomial fit to the coral-based data of Bard et al. (1998). The error in the calendar age is not shown and is assumed to be about the same as the error in the ^{14}C age.

Park, in the northern end of the Sawatch Range, was also the site of an ice-marginal lake that was dammed by a Pinedale glacier (Rosenbaum and Larson, 1983). Lake-bottom sediments from this site, located 0.9 km upstream from the Pinedale terminal moraine, date to 23.2 ± 1.0 ^{14}C (27.4 cal) kyr BP (Fig. 1, Table 1).

Two sites from terminal-moraine complexes indicate that the lag times between the retreat of Pinedale glaciers and the accumulation of datable organic carbon amounted to a few thousand years. For example, bottom sediment from the Winding River kettle located on the Pinedale terminal moraine (2640 m) in the upper Colorado River valley yielded a date of 13.8 ± 0.8 ^{14}C (16.2 cal) kyr BP (Madole, 1976b). Peat that overlies lake sediment deposited at the Mary Jane site (2882 m) in the Fraser River valley yielded a date of 13.74 ± 0.16 ^{14}C (16.13 cal) kyr BP (Nelson et al., 1979).

Pinedale glaciers probably disappeared from the Front Range sometime after 13.7 ^{14}C (16.1 cal) and before 11.8 ^{14}C (13.8 cal) kyr BP. The older limit is based on a date of 13.68 ± 0.11 ^{14}C (16.05 cal) kyr BP for the deglaciation of an area located 1 km west of Buffalo Pass (Madole, 1986), which was near the divide mantled by the Pinedale icecap. The younger limit is based on a lake-sediment ^{14}C age of 11.76 ± 0.10 ^{14}C (13.77 cal) kyr BP from a site located on the Continental Divide ~1.5 km south of Buffalo Pass (Madole, 1980) (Fig. 1, Table 1). La Poudre Pass, a site on the divide of a transection glacier (the largest glacier in the Front Range) that flowed north into the Cache la Poudre River valley and south into the Colorado River valley, was completely ice free by ~10.0 ^{14}C (11.6 cal) kyr BP (Fig. 1, Table 1).

Basal sediments from Front Range cirques have yielded dates indicating the disappearance of Pinedale glaciers prior to 12.0 ^{14}C (14.0 cal) kyr BP. Davis et al. (1992) reported a date of 12.28 ± 0.34 ^{14}C (14.4 cal) kyr from basal sediment in Blue Lake, a high-elevation (3450 m) body of water located at the head of the South St. Vrain Creek drainage basin (Fig. 1, Table 1). In addition, Menounos and Reasoner (1997) obtained a date of 12.04 ± 0.06 ^{14}C (14.0 cal) kyr BP on basal sediment from Sky Pond, another high-elevation (3320 m) lake located in the upper part of the Glacier Creek drainage basin (Fig. 1, Table 1).

Additional dates on basal and near-basal sediment from lakes and fens from relatively high-elevation (>3000 m) sites in north-central Colorado fall within a 10.4 ^{14}C (12.2 cal) to 12.3 ^{14}C (14.4 cal) kyr BP age range (Table 1). We believe that all dates on lake and fen sediments lag the retreat of the Pinedale glaciation by several hundred to a few thousand years, the time it took for abundant vegetation (the source of ^{14}C-datable organic carbon) to be reestablished in surrounding watersheds. ^{14}C dates on peat and beetle fragments should be considered reliable; however, Reasoner and Jodry (2000) found that two dates on bulk organic sediment from Black Mountain Lake in southern Colorado were 760 and 875 yr older than terrestrial macrofossils from the same levels. Thus, the establishment times for some of the lakes and fens may be somewhat later than that estimated using ^{14}C dates obtained on organic carbon from their basal sediments.

To summarize, ^{14}C ages of near-basal sediments from lakes and ponds that formed on drift or that resulted from the damming of ice-marginal drainages indicate (1) formation of Pinedale glaciers was initiated $\geqslant 30$ ^{14}C (35 cal) kyr ago; (2) they were nearing their maximum extent ~22 ^{14}C (26 cal) kyr ago; and (3) they had disappeared from all but the highest elevations of the Front Range by ~14 ^{14}C (16.4 cal) kyr ago.

4. Methods

4.1. Sample collection

Samples were collected with hammer and chisel from tops of boulders located on or near moraine crests. The boulders are Middle Proterozoic granite and quartz monzonite, and ranged in height from 0.3 to 3.0 m (Table 2). The latitude, longitude, and elevation of each boulder was estimated using a hand-held GPS unit. The elevation estimate was further refined using sample locations plotted on topographic maps. The estimated horizontal accuracy was 3–5 m and the estimated vertical accuracy was better than 10 m.

4.2. ^{36}Cl determinations

Samples for ^{36}Cl measurements were sawn to a thickness of 2.5–3.0 cm. Lichens and other organic matter were removed from the surface before reducing the sample to <1 mm fragments by grinding. The sample was then sieved to obtain the >150 μm to <1 mm size fraction. The sample was leached in 3% HNO_3 overnight to remove any secondary chloride and carbonate, rinsed in 18 MΩ water, brought to a basic pH using 1% NaOH, and rinsed several times in 18 MΩ water bringing it to a pH<7. After oven drying at <90°C, aliquots for chemical analyses were removed and ground to a powder using a tungsten-carbide mill.

Major elements, including U and Th, were determined by X-ray fluorescence, and B and Gd were measured by prompt-gamma-emission spectrometry (Table 3). The chloride concentration of the samples was initially estimated with an ion-selective electrode in a Teflon diffusion cell. This initial estimate was used to determine the mass of sample to be dissolved and the amount of stable chloride carrier (99.5% ^{35}Cl) to be added. The chloride concentrations used for age calculations were

Table 2
Heights, snow depth and shielding, locations, elevations, and ^{36}Cl and ^{10}Be age estimations for boulders on moraine crests in the Middle Boulder Creek (MBC), North St. Vrain (NSV) and Roaring Fork (RF) watersheds. FP and JS production refers to ages obtained after application of ^{36}Cl production rates of Phillips et al. (1996, 2001), Stone et al. (1996) and Evans et al. (1997). Erosion rates refer to those rates necessary to increase the apparent of age of sample to 140 kyr

Sample	Moraine type	Boulder height (m)	Hist. Snow mean depth (m)	Snow shielding value	Lat.	Long.	Ele.(m)
MBC-01-1	Pinedale	0.6	0.53	0.922	39.9499	105.5244	2680
MBC-01-2	Pinedale	0.8	0.53	0.930	39.9492	105.5248	2660
MBC-01-3	Pre-Pinedale	0.3	0.53	0.931	39.9479	105.5201	2690
NSV-01-1	Pre-Pinedale	1	0.36	0.990	40.2089	105.5323	2650
NSV-01-3	Pre-Pinedale	3	0.36	1.000	40.2075	105.5219	2610
NSV-01-3dupe	Pre-Pinedale	3	0.36	1.000	40.2075	105.5219	2610
NSV-01-4	Pinedale	1.1	0.36	0.977	40.2235	105.5262	2580
RF-01-2	Pinedale	1.4	0.97	0.881	40.7118	106.5534	2600
RF-01-3	Pinedale	1.2	0.97	0.875	40.7113	106.5534	2620
RF-01-5	Pinedale	1.5	0.97	0.885	40.7097	106.5562	2640
RF-01-6	Pinedale	0.9	0.97	0.865	40.7098	106.5564	2640
RF-01-7	Pinedale	0.5	0.97	0.853	40.7059	106.5588	2660
RF-01-9	Pinedale	1.6	0.97	0.888	40.7055	106.5588	2660
RF-01-10	Pre-Pinedale	1.2	0.97	0.909	40.7142	106.5526	2620
RF-01-11	Pre-Pinedale	1.5	0.97	0.921	40.7149	106.5507	2620
		^{36}Cl ages (kyr)					^{10}Be ages (kyr)
	^{36}Cl/10^{35}Cl	0.0 mm kyr^{-1} erosion, FP production	0.5 mm/ 1.0 mm kyr^{-1} erosion, FP production	Erosion corrected ages Rate/age (mm kyr^{-1}/kyr)	Snow shield corrected	0.0 mm kyr^{-1} erosion, JS production	0.0 mm kyr^{-1} erosion
MBC-01-1	384.4	17.5	17.0/16.5		19.1	18.5	
MBC-01-2	486.7	20.9	20.5/20.1		22.5	20.1	
MBC-01-3	1556	144			157	143	
NSV-01-1	2701	86.9		10.7/140	88		
NSV-01-3	2512	127		6.6/140	127		
NSV-01-3dupe	2493	127			127		
NSV-01-4	882.7	18.4	18.3/18.3		18.9		19.7 ± 1.3
RF-01-2	692.6	25.8	25.3/24.8		28.1[a]		23.2 ± 1.3
RF-01-3	702.1	19.3	19.0/18.8		22.1		
RF-01-5	643.9	19.3	19.0/18.9		21.8		20.1 ± 1.1
RF-01-6	716.5	16.5	16.4/16.3		19.1		
RF-01-7	673.4	17.2	17.0/16.9		20.2		
RF-01-9	497.5	18.4	18.2/17.9		20.8		18.4 ± 1.0
RF-01-10	2954	92.6			103		92.8 ± 5.7
RF-01-11	4062	117		5.9/140	128		

[a] Shielding correction was added to the mean of the ^{36}Cl and ^{10}Be ages.

determined by accelerator mass spectrometry (AMS) using isotope-dilution mass spectrometry.

To extract Cl for ^{36}Cl analyses, a 5:1 mixture of HF and HNO_3 was added to Teflon bottles, containing 20–200 g (depending on initial Cl concentration) of leached sample. The bottles were allowed to sit on a warm hotplate for approximately 5 days to assure complete dissolution. After centrifugation, the supernatant was poured into Teflon beakers and $AgNO_3$ was added to precipitate AgCl. Sulfur removal (^{36}S is an isobar of ^{36}Cl) was accomplished by redissolving the AgCl with NH_4OH and adding $Ba(NO_3)_2$ to precipitate $BaSO_4$. The sample was centrifuged, the solution placed in a clean centrifuge tube, and AgCl was reprecipitated by adding HNO_3. The samples were brought to neutral pH by centrifuging and rinsing in 18 MΩ and dried on covered watch glasses in an oven at $<70°C$.

$^{36}Cl/^{37}Cl$ and $^{35}Cl/^{37}Cl$ ratios of the purified AgCl precipitate were measured at the Center of Accelerator Mass Spectrometry facility at the Lawrence Berkeley National Laboratory under the supervision of R. Finkel. ^{36}Cl surface-exposure ages were calculated using the most recent version of the CHLOE program (Phillips and Plummer, 1996). Muon production in CHLOE is based on the work of Evans et al. (1998). In CHLOE the calculation of low-energy neutron absorption, the

Table 3
Chemical compositions of boulders sampled on moraine crests in the Middle Boulder Creek (MBC), North St. Vrain (NSV) and Roaring Fork (RF) watersheds

Sample	SiO_2 (wt%)	TiO_2 (wt%)	Al_2O_3 (wt%)	Fe_2O_3-T (wt%)	MnO (wt%)	MgO (wt%)	CaO (wt%)	K_2O (wt%)	Na_2O (wt%)	P_2O_5 (wt%)	LOI (wt%)
MBC-01-1	74.96	0.09	14.41	1.60	0.01	0.30	3.31	0.51	4.58	0.01	0.25
MBC-01-2	72.22	0.40	14.06	2.82	0.03	0.48	1.03	5.34	3.03	0.04	0.53
MBC-01-3	64.96	0.99	16.24	5.94	0.07	1.25	2.68	3.44	3.79	0.13	0.37
NSV-01-1	75.33	0.32	12.78	2.01	0.02	0.40	0.83	5.35	2.58	0.10	0.31
NSV-01-3	70.94	0.48	14.65	5.22	0.05	1.04	0.66	5.00	2.01	0.10	0.23
NSV-01-4	76.90	0.18	12.84	1.08	0.01	0.23	0.65	5.67	2.70	0.11	0.33
RF-01-2	76.11	0.16	13.21	1.27	0.03	0.21	1.12	4.50	3.44	0.01	0.39
RF-01-3	76.19	0.17	13.38	1.32	0.03	0.21	1.02	4.50	3.53	0.02	0.37
RF-01-5	77.05	0.12	12.77	1.21	0.03	0.18	0.88	4.41	3.44	0.02	0.30
RF-01-6	75.21	0.21	13.13	1.99	0.04	0.30	1.03	4.50	3.43	0.02	0.50
RF-01-7	78.59	0.09	12.65	0.46	0.01	0.09	0.51	3.93	4.11	0.02	0.37
RF-01-9	74.25	0.29	13.76	2.14	0.04	0.27	1.43	3.47	4.05	0.02	0.33
RF-01-10	76.03	0.11	13.04	1.14	0.02	0.13	0.87	4.56	3.49	0.03	0.33
RF-01-11	80.41	0.04	11.22	0.53	0.00	0.03	0.18	4.47	3.52	0.01	0.16
	Cl (ppm)	Ba (ppm)	Rb (ppm)	Sr (ppm)	Pb (ppm)	Th (ppm)	U (ppm)	B (ppm)	Gd (ppm)		Total (wt%)
MBC-01-1	101	56	25	343	13	12.6	11	<3	2		100.0
MBC-01-2	123	1020	217	270	33	36.7	2	<3	4		100.0
MBC-01-3	325	1130	180	426	27	59.4	3	<3	10		99.9
NSV-01-1	106	416	314	103	35	23.9	3	<3	4		100.0
NSV-01-3	159	575	197	132	37	12.9	3	<3	5		100.4
NSV-01-4	49	318	244	68	32	6.6	1	<3	2		100.7
RF-01-2	89	826	167	161	21	4.4	3	<3	1		100.5
RF-01-3	78	711	196	144	27	16.7	5	<3	2		100.7
RF-01-5	64	560	210	112	24	2.8	3	<3	2		100.4
RF-01-6	50	880	189	170	21	20.1	3	<3	4		100.4
RF-01-7	52	646	134	139	14	11.4	6	3	1		100.8
RF-01-9	76	984	156	249	19	6	2	<3	3		100.0
RF-01-10	79	841	184	162	24	9.2	2	<3	3		99.8
RF-01-11	53	136	205	23	17	7.1	9	<3	3		100.6

quantification of the flux distribution, and its dependence on chemical composition is based on work discussed in Phillips et al. (2001). Spallation production constants for ^{36}Cl from Ca and K (66.8 and 154 atoms g^{-1} yr^{-1}) derived by Phillips et al. (1996, 2001) were used in all age calculations. In addition, spallation production constants for ^{36}Cl from Ca and K (48.8 and 170 atoms g^{-1} yr^{-1}) derived by Stone et al (1996) and Evans et al. (1997) were applied to calculation of the ^{36}Cl ages of three Middle Boulder Creek samples. All ages were calculated using a bulk density of 2.7 g cm^{-3} and a neutron attenuation coefficient of 170 g cm^{-2}. The analytical error for ^{36}Cl ages is assumed to be $\leqslant 8\%$ of the age (Phillips et al., 1997). The 8% value is probably an overestimate of analytical error in this study; i.e., the reproducibility of two independent analyses of sample NSV-01-3 was excellent with no difference in age at the 3rd significant figure (Table 2). Chlorine analytical data are listed in Appendix A of the electronic supplement to this paper.

4.3. ^{10}Be determinations

Samples for ^{10}Be measurements were sawn to a uniform thickness (4–5 cm), crushed, and quartz concentrated from 250 to 500 μm fraction using magnetic and heavy-liquid techniques. All samples were treated in >1 M HCl on shaker table for 24 h to remove carbonates and oxide coatings on quartz. Samples were then rinsed in de-ionized water and air-dried. Pure quartz separates were prepared by etching concentrates in a solution of 2% HF and 1% HNO_3 over a 5-day period in an ultrasonic bath. The concentrate-to-solution ratio was 7 g/l. Acids were changed daily. This process yielded nearly pure quartz with trace amounts of a metallic sulfide that were removed by hand.

From 20 to 30 g of quartz separate were dissolved in HF and spiked with 300 μg of Be carrier (Spectrosol Be standard 1000 mg/l with a $^{10}Be/^{9}Be$ ratio of 2.3×10^{-14}). A procedural blank consisting of pure carrier was run at the same time as the samples. Be was separated from other species (chiefly Fe, Al, and Ti) using ion-exchange columns and precipitation of $Be(OH)_2$ at a pH of 8.5–9.0. The hydroxide was converted to BeO by firing in a quartz crucible over an open flame until incandescent. Powdered Ag was mixed with BeO in the crucible and pressed into a cathode for analysis by AMS.

AMS analyses were conducted at the PSI/ETH AMS facility with a terminal voltage of 5.8 MV. The reported isotope ratios were normalized to Be standard S555 with a nominal value of $^{10}Be/Be = 95.5 \times 10^{-12}$. All ages were corrected for the carrier blank and for sample thickness using a bulk density of $2.7\,g\,cm^{-3}$ and a neutron attenuation coefficient of $170\,g\,cm^{-2}$.

Exposure ages were computed using a ^{10}Be production rate of $5.37 \pm 0.22\,atom\,g^{-1}\,yr^{-1}$ at sea level and high latitude (SLHL). This production rate is from calibration sites in Austria (Kubik et al., 1998; Schaller et al., 2001) and Wyoming (Klein and Gosse, 2002) that are the best available matches for the latitude, elevation, and exposure time of our Colorado sites described in this paper. Production rates were scaled from SLHL to our Colorado sites using the procedures of Stone (2000), with the fraction of production by neutron spallation set at 0.974. Errors (1σ) are analytical (AMS and target chemistry) uncertainties plus the error reported for the SLHL production rate.

No correction was made of dipole-induced variability of production rates. Geographic latitude was used in scaling of Colorado production rates in that production rates at mid-latitudes are not thought to be sensitive to geomagnetic field strength (Cerling and Craig, 1994; Phillips et al., 1996). Beryllium analytical data have been listed in Appendix B of the electronic supplement to this paper.

4.4. Snow-shielding calculations

To examine the effect of snow shielding on the apparent (measured) ages of Pinedale and pre-Pinedale boulders, the mean snow cover for the past ~25 kyr and ~140 kyr must be estimated. One way to approximate changes in snow cover is to determine changes in effective wetness (W_{eff}) for a surface-water system that drains a mountainous area. Unfortunately paleohydrologic records of W_{eff} are generally not available for much of the western US, but tend to be concentrated in the Great Basin (e.g., Oviatt et al., 1992; Benson et al., 1995). Obviously, changes in W_{eff} of the Rockies of north-central Colorado have not necessarily paralleled changes in W_{eff} of the Sierras or the Wasatch Mountains over the past 25 or 140 kyr. However, to illustrate the hypothetical importance of snow shielding on Colorado moraine-crest boulder ages we applied relative changes in W_{eff} that occurred in the Sierra Nevada (as measured by changes in the hydrologic balances of Lake Lahontan and Owens Lake; Benson and Peterman, 1995; Benson et al., 1995, 2002; Menking et al., 1997; Benson, 2003) to the north-central Colorado. The boulder ages obtained after correction for snow shielding should be considered only as rough approximations given that snow-cover data are not available for our Colorado sites during the Pleistocene and Holocene.

The mean effective wetness ($\bar{W}_{eff}$) for the Sierra Nevada during the past 25 kyr was determined from direct sources (elevations of ^{14}C-dated lacustrine carbonates) and indirect sources (^{14}C-dated core-based $\delta^{18}O$ lacustrine records) of lake-level data compiled by Benson and Peterman (1995) and Benson et al. (1995, 2002). The 25-kyr record was extended to 140 kyr by splicing the hydrologic-balance record from Owens Lake (Menking et al., 1997; Benson, 2003) to the Pyramid Lake record. $\bar{W}_{eff}$ values of 4.3 and 3.3 were calculated for the past 25 and 140 kyr (Fig. 3).

In order to apply Sierran values of $\bar{W}_{eff}$ to Colorado sites, we scaled historical snow-depth data from Colorado snow-course sites located at elevations and topographic settings similar to our sampling sites (e.g., the Willow Creek Pass site shown in Fig. 4) using a scaling factor of 4.3 or 3.3. Snow-shielding values were calculated for all north-central Colorado boulders that experienced such hypothetical increases in snow cover. We used a snow density of $0.27\,g\,cm^{-3}$ and the following equation (Gosse and Phillips, 2001) to calculate the fraction of the cosmic-ray flux reaching the rock surface that lay beneath the snow cover:

$$S_{snow\ on\ rx} = \frac{1}{12}\sum_{i=1}^{i=12} e^{-([D_{snow,i} - D_{rx}]\rho_{snow,i})/A_f},$$

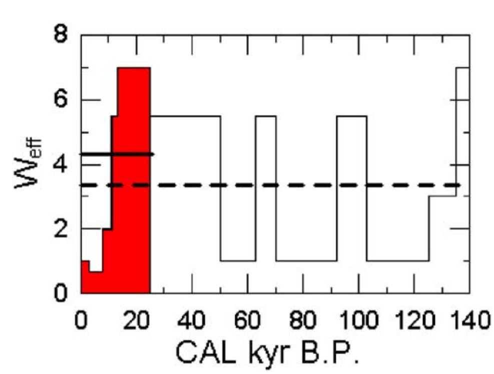

Fig. 3. Estimated effective wetness (relative to historical pristine reconstructions) for the northern Sierra Nevada, California. Shaded area indicates the history of effective wetness experienced by the Lake Lahontan area, Nevada and California, for the past 25 kyr. Solid line denotes the mean effective wetness for the past 25 kyr; dashed line denotes the mean effective wetness for the past 140 kyr.

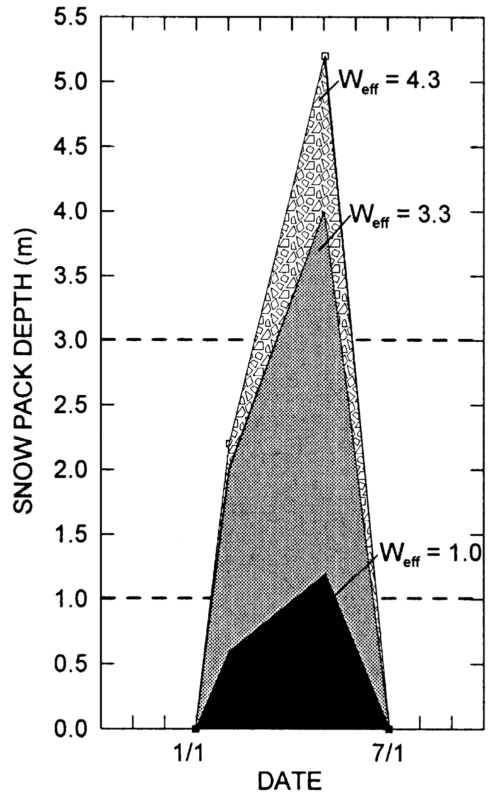

Fig. 4. Willow Creek Pass (elevation 2908 m) snowpack depths scaled to 3.3 × and 4.3 × historical mean values for the months February, March, April, and May. The historical snowpack has a mean depth of 0.94 m.

where $S_{\mathrm{snow\ on\ rx}}$ is the annual snow shielding, $D_{\mathrm{snow},i} - D_{\mathrm{rx}}$ is the monthly amount of snow cover on the rock surface, $\rho_{\mathrm{snow},i}$ is the density of the snow pack ($\sim 0.27\,\mathrm{g\,cm^{-3}}$), and A_{f} is the snow-free cosmic-ray-flux attenuation length.

5. Results

In this section we first compare the ^{36}Cl ages of three boulders from the Middle Boulder Creek area using two different sets of ^{36}Cl production values. We then compare the ^{36}Cl age estimates of five boulders with their ^{10}Be age estimates. Given the complexity of ^{36}Cl production, the ^{10}Be dates are used to demonstrate the relative accuracy of the ^{36}Cl results. After showing that applications of the various age-estimation methods do not result in significant age differences, we use the Phillips et al. (1996, 2001)-based ^{36}Cl ages in calculations of the effects of erosion and snow shielding on the ages of Pinedale and pre-Pinedale boulders.

5.1. Comparison of ^{36}Cl ages based on different production values

The Middle Boulder Creek samples evaluated using the production values of Stone et al. (1996) and Evans et al. (1997) resulted in ^{36}Cl age estimates that differed by $\leqslant 1$ kyr from ^{36}Cl ages based on the Phillips et al. (1996, 2001) production values (Table 2). Therefore, in the following, we use the ^{36}Cl ages that are based on the production values of Phillips et al. (1996, 2001).

5.2. Comparison of ^{36}Cl and ^{10}Be age estimates

Zero-erosion ^{10}Be ages of four Pinedale boulders differ from the Phillips et al. (1996, 2001)-based ^{36}Cl zero-erosion age estimates by 0.0–2.6 kyr (Table 2). The only sample that might be argued to have a ^{10}Be date significantly younger than its ^{36}Cl counterpart is RF-01-2. If so, this would imply that the surface of RF-01-2 has been eroded and that the true date of surface exposure should fall somewhere between 25.8 and 23.2 kyr. However, the fact that the exposure age of this boulder is significantly older than all other Roaring Fork boulder exposure ages suggests that the boulder contains inherited ^{36}Cl and ^{10}Be.

5.3. Comparison of ^{36}Cl ages with continental and Sierran ice records

Five pre-Pinedale boulders have ^{36}Cl ages ranging from 87 to 144 kyr B.P. (Table 2). Only the 144 kyr date (MBC-01-3) falls within a "glacial" period; i.e., marine-isotope stage (MIS) 6 (Martinson et al., 1987). The four other pre-Pinedale boulders (NSV-01-1, RF-01-10, RF-01-11, and NSV-01-3) have surface-exposure ages that fall within MIS 5, a period in which global sea-level records indicate that continental ice sheets were small (Shackleton, 1987) and Sierra Nevada ice fields were small or nonexistent (Bischoff et al., 1997) (Fig. 5). We also plotted the ^{36}Cl ages of Bull Lake boulders from the Wind River Range, Wyoming (Phillips et al., 1997) (Fig. 5B). Only 8 of 30 boulders have surface-exposure ages that fall within MIS 4 or 6.

5.4. Processes leading to anomalous ^{36}Cl ages

There are several reasons why a boulder surface may have a ^{36}Cl age that is younger or older than expected, including inheritance, erosion of the boulder surface, and shielding (by snow or sediment), and ^{36}Cl leakage. Inheritance occurs when a rock fails to lose the cosmogenic signature acquired prior to transport by glacier ice. For example, erosion may not remove all of the inherited ^{36}Cl from a glacial boulder remobilized during a subsequent glacial advance. This results in ^{36}Cl ages that are older than the time of boulder deposition.

Surface erosion can either increase or decrease the ^{36}Cl age of a boulder surface. The apparent ^{36}Cl age of a boulder increases with erosion to a depth of $\sim$15 cm, which is the depth of the production maximum resulting from thermal and epithermal neutron absorption

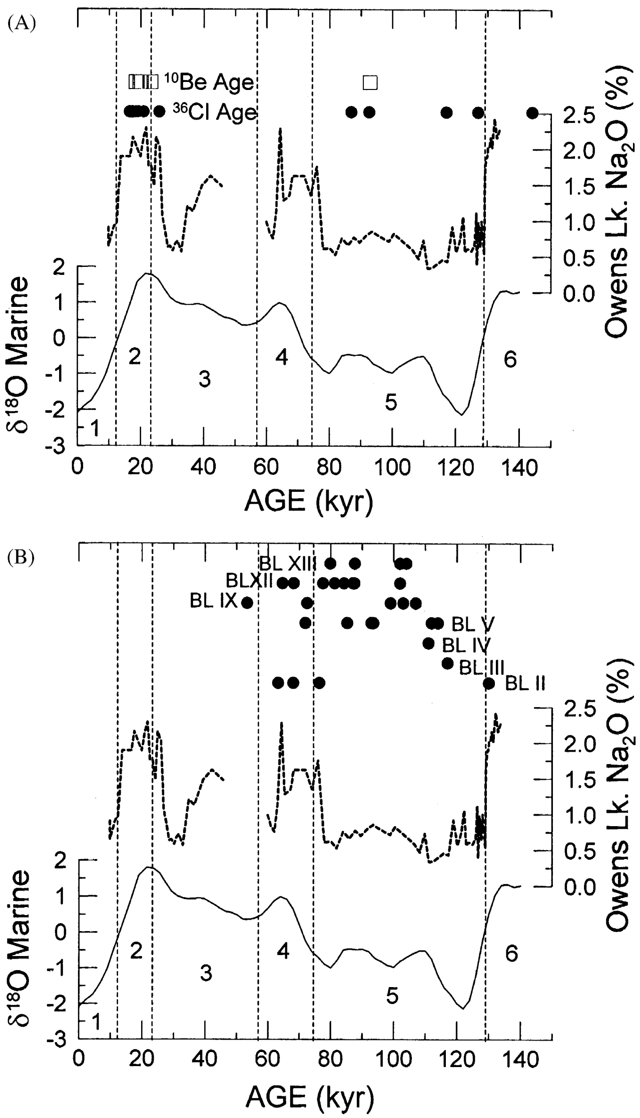

Fig. 5. Uncorrected ^{10}Be and ^{36}Cl ages of boulder surfaces from north-central Colorado (A) and Wyoming (B) plotted together with an Owens Lake proxy indicator (Na_2O) of Sierran alpine glaciation (dashed line) and a marine proxy indicator ($\delta^{18}O$) of continental ice volume. Increasingly positive values of both indicators indicate greater ice volumes. Bull Lake sequences in Wyoming are prefaced with a BL.

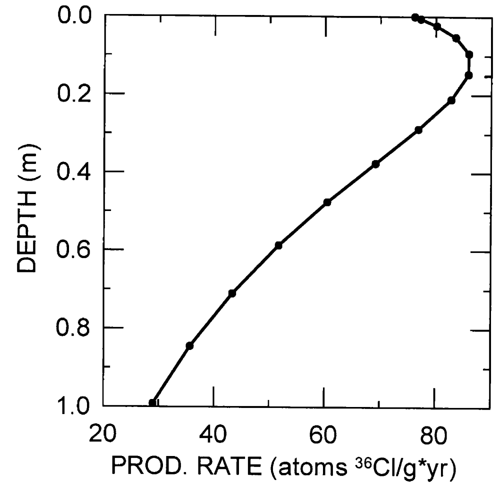

Fig. 6. Simulated production rate plotted as a function of depth below the upper surface of boulder NSV-01-1. Note the maximum in production at ~15 cm. For erosion confined to the upper 30 cm, the apparent ^{36}Cl age will be too old and for erosion below 30 cm, the apparent ^{36}Cl age will be too young.

(Fig. 6). With increasing depth, the apparent age decreases exponentially.[2] For the most part, Pinedale boulders have not undergone severe surface erosion. Caine (1979) monitored the weathering of crushed rhyodacite in a broad range of microenvironments, above treeline, in the San Juan Mountains of south-western Colorado over a 5-year period. His data indicated a mean rate of rock-surface retreat of 0.00055 mm yr^{-1}. Benedict (1993) measured phenocryst relief on a southwest-facing granodiorite in a col located at an elevation of 3505 m in the northern Front Range of Colorado. For conditions in which present-day snowcover is less than ~30 weeks during the year, maximum pheoncryst relief averaged ~10 mm. If we assume that the Pinedale glacier had disappeared from this area between 12 and 14 cal kyr B.P. (see Section 3 above), the mean rate of rock-surface retreat was somewhere between ~0.7 and ~0.8 mm kyr^{-1}. In order to correct for rock-surface retreat, erosion rates of 0.5 and 1.0 mm kyr^{-1} were used in calculating ^{36}Cl ages. The two erosion rates decreased the apparent ^{36}Cl ages of boulders analyzed in this study by 100–500 yr (0.5 mm kyr^{-1} rate) and by 100–1000 yr (1.0 mm kyr^{-1} rate) (Table 2).

Because some granite boulders may sporadically spall, the effect of rate of spall on the calculated ^{36}Cl age was estimated. A line was fit to the ^{36}Cl production rate (atoms Cl g^{-1}yr^{-1}) for RF-01-2 and the fit line was used to increment the production values at 0.1- and 1-mm depth intervals. The ^{36}Cl inventory of a sample with an apparent age of 25 kyr (value calculated assuming zero erosion) was created by multiplying the annual production rate by 25,000. Erosion simulations were conducted at a mean rate of 1 mm kyr^{-1}, assuming three different rates of spall (0.1 mm 100 yr^{-1}, 1 mm kyr^{-1}, and 5 mm 5 kyr^{-1}). The lowest rate of spall is thought to approximate continuous erosion. The sample age (integrated production over the top 25 mm of sample) decreased slightly as a function of spall rate; i.e., from 26.07 to 25.90 to 25.75 kyr for the three rates, suggesting that the assumption of continuous erosion does not greatly affect the calculated age of a Pinedale sample that may have eroded discontinuously by spalling.

It appears improbable, during MIS 5 (see Section 5.3), that valley glaciers were large enough to transport

[2] Erosion down to a depth of ~30 cm in NSV-01-1 will produce apparent ^{36}Cl ages that are too old.

boulders NSV-01-1, NSV-01-3, and RF-01-11 to their present locations: therefore, the erosion rate in CHLOE was varied in order to shift the apparent ^{36}Cl ages of these boulders to MIS 4 or MIS 6. Sharp et al. (2003) have recently shown that the glaciofluvial terrace WR3 in the Wind River Basin, which formed penecontemporaneously with the final maximum Bull Lake glacial advance, has a ^{230}Th/U age of 150 ± 8.3 kyr age when corrected for incision. This finding suggests that the Bull lake glaciation was broadly synchronous with the penultimate global ice volume maximum during MIS 6 and did not occur during MIS 5.

Dates of 60 and 140 kyr were arbitrarily selected as the times the boulders would have been deposited during MIS 4 or MIS 6. MIS 4 ages for the samples could not be obtained by varying the rate of erosion; however, calculations indicated that erosion rates of 5.9, 6.6, and 10.7 mm kyr^{-1} could reduce the estimated ages of 140-kyr-old boulders to the observed values (87, 127, and 117 kyr, Table 2). The calculated erosion rates seem excessive given that Pinedale boulders selected for cosmogenic dating generally appear to have been eroded at rates $\leqslant 1$ mm kyr^{-1}. Even though physical weathering of boulders may have been more rapid between 15 and 75 kyr, when the snowpack was estimated to be much thicker than today, it is difficult to call on such rapid rates of weathering to explain these anomalous ages; i.e., these rates of weathering would have removed 0.8–1.5 m from the surfaces of these boulders during the past 140 kyr. In fact, Dahl (1967) showed that granite bedrock surfaces that were snow covered for most of the year showed much less weathering than other surfaces at the same elevation. Benedict (1993) also showed that granodiorite weathering is least effective beneath late-lying snowbanks that protect the rock from freeze-thaw and wetting-drying cycles. Their data argue for lower weathering rates during the somewhat wetter MIS 5 period.

Pre-Pinedale samples were much easier to trim with a diamond saw than Pinedale samples. In addition, the pre-Pinedale samples tended to break apart when sawing, suggesting that they possess a significant amount of secondary porosity. We offer the hypothesis that Cl leaching may have accompanied and followed the development of porosity. If ^{36}Cl has been preferentially leached from a sample, relative to the Ca and K from which the ^{36}Cl formed, the ^{36}Cl age would underestimate the original time of exposure.

The above discussion suggests that other processes, such as snow shielding, sediment shielding, and ^{36}Cl leakage may have been responsible for the relatively young surface ages of NSV-01-1, NSV-01-3, and RF-01-11. The young ^{36}Cl age of RF-01-10 ($\sim$93 kyr) may also be due to some combination of snow and sediment shielding. But the age cannot be explained by ^{36}Cl leakage or rock-surface erosion because the sample possesses a nearly identical ^{10}Be age (Table 2).

As mentioned previously, pre-Pinedale end moraines have lower and broader profiles than Pinedale end moraines. The low profiles of pre-Pinedale moraines may be the result of slope erosion or mass movement (Hallet and Putkonen, 1994) and may involve post-depositional disturbance of boulders such as rolling or overturning. The rate of erosion of till matrix is high compared to that of boulder surfaces; therefore, some boulders now resting on pre-Pinedale moraines are likely to have been exhumed since their deposition. In addition, mean snow depths must have been higher during MIS 2, 3, and 4 ($\sim$13 to 75 kyr BP, Fig. 5) than during the past 50 yr.

Whereas it is not possible to directly determine the effects of sediment shielding on the boulders investigated in this paper, approximate snow-shielding values can be applied to all samples assuming changes in $\bar{W}_{\text{eff}}$ discussed in Section 5.4. For most (six) Pinedale boulders, snow-shielding corrections yielded age increases ranging from 1.6 to 3.0 kyr; and for three pre-Pinedale boulders, snow-shielding corrections yielded age increases ranging from 10 to 13 kyr. Only for boulders whose upper surfaces stand well above present-day snowpack (e.g. NSV-01-1, NSV-01-3, and NSV-01-4) was the shielding correction minimal relative to the apparent age of the sample (Table 2). The snow-shielding corrections for smaller pre-Pinedale boulders surfaces are substantial; however, they are not enough by themselves to produce MIS 6 ages. This suggests that soil shielding and perhaps ^{36}Cl leakage has affected most of the pre-Pinedale boulder surfaces examined in this paper; i.e., boulders have emerged from the coarse till in which they were deposited or have rolled since their deposition or exhumation.

6. Summary and conclusions

Uncorrected ^{36}Cl ages of eight Pinedale boulders in north-central Colorado fall in the range 16.5–20.9 kyr (sample RF-01-2, which is believed to have acquired inherited ^{36}Cl, was excluded from this data set). ^{10}Be age determinations on four boulder surfaces are in close agreement with ^{36}Cl determinations, suggesting that surface erosion has been minimal during the past $\sim$22 kyr. Corrections for snow shielding increased the ^{36}Cl ages of Colorado boulders by as much as 3 kyr, with increases in age averaging 2.2 kyr (12% of the apparent age). The magnitude of this correction, although only an approximation, illustrates the importance of snow shielding on boulders whose heights are not significantly greater than the height of the mean historical snowpack. Without accounting for snow shielding on such relatively small boulders, it may prove difficult to match moraine ages with millennial-scale phenomena such as Dansgaard–Oeschger or Heinrich events.

Most ages for Colorado and Wind River, Wyoming, pre-Pinedale (Bull Lake) boulders fall within MIS 5, a time when continental and Sierran ice accumulations were small or nonexistent. Under the assumption that the Colorado boulders were deposited on moraines that formed near the end of MIS 6 (i.e., 140 kyr), calculations indicate that rock-surface erosion rates would have had to range from 5.9 to 10.7 mm kyr^{-1} to produce the apparent ^{36}Cl values present in the top 2–3 cm of these rocks. If correct, these calculations indicated that from 0.8 to 1.5 m of the surfaces of the boulders were removed during the past 140 kyr. However, the calculated erosion rates can be argued to be anomalously rapid when compared to rates that have been documented for the past 20 kyr. In addition, snow shielding only accounts for 0–48% of the additional years needed to shift pre-Pinedale dates to MIS 6. This suggests that sediment shielding and perhaps ^{36}Cl leakage has decreased the apparent surface ages of most pre-Pinedale boulders described in this paper. Given that the rate of erosion of till matrix is rapid compared to that of boulder surfaces, it is probable that the upper surfaces of many boulders found on pre-Pinedale moraines were exposed to cosmic radiation some time after they were deposited. The inability to determine the magnitudes of snow and sediment shielding that affected pre-Pinedale boulder surfaces thus creates significant uncertainty about using their apparent cosmogenic ages as accurate estimators of the timing of middle-Pleistocene alpine glaciation.

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Appendix A. Chlorine analytical data

Cl Analytical are presented in Table 4.

Appendix B. Beryllium analytical data

Be analytical data are presented in Table 5.

Table 4

NAME	CAMS #	S FOM	^{36}Cl/^{37}Cl ratio		Background		Bkgd-corrected		^{35}Cl/^{37}Cl ratio	
			Ratio	±	Ratio	±	Ratio	±	Ratio	±
MBC-01-1	CL10160	0.999	2.598E−12	6.11E−14	8.21E−14	1.23E−14	2.518E−12	6.24E−14	5.55	0.001
MBC-01-2	CL10161	0.999	3.375E−12	7.93E−14	8.21E−14	1.23E−14	3.295E−12	8.03E−14	5.77	0.004
MBC-01-3	CL10162	1.000	1.042E−11	1.81E−13	8.21E−14	1.23E−14	1.035E−11	1.82E−13	5.65	0.007
NSV-01-1	CL10163	1.000	1.433E−11	2.49E−13	8.21E−14	1.23E−14	1.426E−11	2.49E−13	4.28	0.001
NSV-01-2	CL10164	0.999	6.578E−12	1.44E−13	8.21E−14	1.23E−14	6.500E−12	1.44E−13	5.75	0.001
NSV-01-3	CL10165	1.000	1.435E−11	2.49E−13	8.21E−14	1.23E−14	1.427E−11	2.49E−13	4.68	0.000
NSV-01-3d	CL10166	1.000	1.414E−11	2.22E−13	8.21E−14	1.23E−14	1.406E−11	2.22E−13	4.64	0.003
NSV-01-4	CL10167	0.999	5.842E−12	1.12E−13	8.21E−14	1.23E−14	5.764E−12	1.13E−13	5.53	0.011
RF-01-2	CL10169	0.999	4.821E−12	1.12E−13	8.21E−14	1.23E−14	4.742E−12	1.13E−13	5.89	0.004
RF-01-3	CL10170	0.999	4.095E−12	1.06E−13	8.21E−14	1.23E−14	4.016E−12	1.06E−13	4.72	0.003
RF-01-5	CL10171	0.999	4.496E−12	1.24E−13	8.21E−14	1.23E−14	4.417E−12	1.25E−13	5.86	0.009
RF-01-6	CL10172	0.999	4.729E−12	1.76E−13	8.21E−14	1.23E−14	4.650E−12	1.76E−13	5.49	0.010
RF-01-7	CL10173	0.999	4.456E−12	1.52E−13	8.21E−14	1.23E−14	4.377E−12	1.53E−13	5.50	0.020
RF-01-9	CL10174	0.999	3.473E−12	9.37E−14	8.21E−14	1.23E−14	3.393E−12	9.45E−14	5.82	0.010
RF-01-10	CL10175	1.000	1.689E−11	6.08E−13	8.21E−14	1.23E−14	1.681E−11	6.08E−13	4.69	0.005
RF-01-11	CL10176	1.000	2.642E−11	6.20E−13	8.21E−14	1.23E−14	2.636E−11	6.20E−13	5.49	0.001
AgCl Blank	CL10158	0.955	8.204E−14	1.230E−14			8.209E−14	1.230E−14	3.16	0.003

Note: 1. ^{36}Cl/Cl ratios were normalized to a standard (KNSTD) with a ^{36}Cl/Cl ratio of 1.60E−12.

2. Ratios were corrected for spurious counts resulting from sulfur contamination in the sample. A sulfur figure of merit of 1 indicates that there were no spurious counts, an FOM of .95 corresponds to a 5% sulfur correction, etc. The FOM does not correct for the count rate suppression that can occur at high *S* rates.

3. Background-corrected ratios were calculated using the backgrounds indicated, based on measured blank CL10158.

4. ^{35}Cl/^{37}Cl ratios were measured relative to standards with an assumed natural 35/37 ratio of 3.127. The uncertainty quoted reflects the standard deviation of the multiple measurements for each sample.

Table 5

Sample ID	AMS ID	$^{10}Be/^{9}Be$ ($\times 10^{-12}$)	Error (%)	Be carrier (mg)	Sample mass (g)	$^{10}Be^{a}$ (atoms/g $\times 10^4$)	Thickness (cm)	Scaling factor[b]
NSV-1-4	ZB2233	0.7681	4.80	0.3559	26.021	67.06	4.0	6.61
RF-1-2	ZB2234	0.6311	4.00	0.3575	17.940	79.46	5.4	6.74
RF-1-5	ZB2235	1.1700	4.00	0.3573	38.050	71.26	4.0	6.89
RF-1-9	ZB2236	0.7141	4.00	0.3579	24.734	65.72	4.4	6.95
RF-1-10	ZB2237	3.8029	4.70	0.3566	27.898	321.88	2.4	6.77

The AMS results have been normalized to Zurich ETH standard S555 with a nominal value of $^{10}Be/Be = 95.5 \times 10^{-12}$.
This is a secondary standard normailzed to the material BEST433 (H. Hofmann et al, Nucl. Instr. and Meth. B29 (1987) 32–36).
The uncertainty of this standard (about 2.5%) is not included in the quoted uncertainties of the samples.
Quoted error is $1\,\sigma$.
[a] Value corrected for beryllium procedural blank value of $0.0344 \times 10–12 \pm 15\%$.
[b] Production rate scaling factor computed after Stone (2000) for sample geographic latitude and altitude.

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